National Aeronautics and Space Administration
Washington 25, D.C.

(NASA CR - - -) OTS:

Covering the period 6 October 1963 thru 5 February 1964

GEOPHYSICS CORPORATION OF AMERICA
Bedford, Massachusetts

A STUDY OF THE METEOROLOGY OF
MARS AND VENUS

Quarterly Progress Report No. 4

(NASA Contract No. NASw-704)

auth [1964] 25 p refs

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I. INTRODUCTION

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During the past quarter, no new investigations were begun, but work started in previous quarters was continued. Computations of the radiative equilibrium distribution of temperature between the cloud-base and the surface of Venus were completed, and the results are presented in this progress report. To the extent that this region of the Venus atmosphere is in radiative equilibrium, and to the extent that the computed temperature distribution is thermally stable, the results provide estimates of the actual temperature versus altitude distribution below the clouds.

Work has continued on the construction of a model radiation budget for the planet Mars. In the present report, we describe our progress in this area and present a preliminary version of the average annual radiation budget at the top of the Martian atmosphere.

Thermodynamic diagrams for the Martian atmosphere, as described in Quarterly Progress Report No. 3, are being constructed and will be available for the Final Report. Author

II. METEOROLOGY OF VENUS

The model used for the computation of the radiative equilibrium distribution of temperature between the cloud-base and the surface of Venus has been described in Quarterly Progress Report No. 3. Very briefly, it is assumed that the cloud layer completely covers the sky; both the cloud-base and surface radiate as black bodies; the atmospheric layer below the cloud is a grey absorber and is in infrared radiative equilibrium; and both the cloud-base and surface are maintained at fixed temperatures, which are chosen to be in agreement with present indications of these quantities.

The surface temperature is taken as 700°K, based upon the radio-wavelength emission observations. The cloud-base temperature is taken as 373°K (100°C) based upon the value of about 200°F (366°K) given in the report, Mariner: Mission to Venus. This value is only an estimate, since the cloud-base temperature has never been measured. The only other parameter that must be specified before computations can be performed is the infrared opacity of the atmosphere between the surface and the cloud-base. Since this parameter is unknown, we have performed computations for a range of τ_b values, including $\tau_b = 1, 2, 3, 5$, and 10. The atmospheric layer is broken up into 15 intervals, and temperatures are computed for each interval. The results are presented in Table 1. In this table the τ values are mid-points of the layers for which the temperatures were computed.

Table 1. Radiative equilibrium temperatures below the cloud on Venus.

	$\tau_b = 1$		$\tau_b = 2$		$\tau_b = 3$		$\tau_b = 5$		$\tau_b = 10$	
	τ	$T(^oK)$	τ	$T(^oK)$	τ	$T(^oK)$	τ	$T(^oK)$	τ	$T(^oK)$
Ground	0	700	0	700	0	700	0	700	0	700
	.03	653	.07	665	.10	672	.17	678	.33	685
	.10	645	.20	656	.30	662	.50	668	1.00	674
	.17	637	.33	647	.50	652	.83	658	1.67	663
	.23	630	.47	638	.70	643	1.17	647	2.33	652
	.30	623	.60	629	.90	633	1.50	636	3.00	640
	.37	615	.73	620	1.10	622	1.83	625	3.67	628
	.43	608	.87	610	1.30	612	2.17	613	4.33	614
	.50	600	1.00	600	1.50	600	2.50	600	5.00	600
	.57	592	1.13	590	1.70	588	2.83	586	5.67	585
	.63	584	1.27	578	1.90	575	3.17	572	6.33	568
	.70	575	1.40	566	2.10	561	3.50	556	7.00	550
	.77	565	1.53	553	2.30	546	3.83	538	7.67	530
	.83	554	1.67	539	2.50	529	4.17	518	8.33	507
	.90	543	1.80	522	2.70	510	4.50	496	9.00	480
	.97	529	1.93	503	2.90	487	4.83	468	9.67	447
Cloud-base	1.00	373	2.00	373	3.00	373	5.00	373	10.00	373

For a grey atmosphere, the infrared opacity is proportional to the pressure, and we can determine the temperature distribution as a function of atmospheric pressure. We take the value $p = 2 \times 10^{-1}$ atm for the pressure at the cloud base; this value is based upon our temperature of 373°K and the temperature - pressure model atmosphere presented by Kaplan (1963). For the pressure at the surface, we take the value 10 atm, again based upon Kaplan's (1963) model.

For a grey atmosphere, we have

$$d\tau = \alpha \rho \, dz = - \frac{\alpha}{g} dp$$

where τ is infrared opacity, α is the grey absorption coefficient of the atmosphere, ρ is density of the atmosphere, z is height, and p is pressure. Therefore,

$$d\tau = - A \, dp$$

or

$$\tau = - A \, p + B$$

where $A = \left(\frac{\alpha}{g} \right)$ and B is a constant. A and B can be determined from the conditions $\tau = \tau_b$ at $p = 2 \times 10^{-1}$ atm, and $\tau = 0$ at $p = 10$ atm. The resulting relationship is

$$p = \frac{10 \, \tau_b - 9.8 \, \tau}{\tau_b} .$$

This permits conversion of opacity (τ) values to pressures (p). We have plotted the computed temperatures versus pressure on a semi-log scale, and these distributions are shown as Figures 1 to 5. The semi-log scale is approximately equal to a linear height scale. On each diagram an adiabatic curve is also plotted from the relationship

$$\left(\frac{T}{T_o}\right) = \left(\frac{p}{p_o}\right)^{0.286}$$

In general, the computed temperature curves are characteristically super-adiabatic in the atmospheric layer closest to the planet's surface and in the layer immediately below the cloud top. This is not surprising, for previous radiative equilibrium computations have shown that there is a discontinuity in temperature between the surface and atmosphere at the boundary of a layer in radiative equilibrium (See, for example, Goody, 1954). As the infrared opacity, τ_b, increases, the depth of the layers with super-adiabatic lapse rates decreases, and the lapse rates in these layers approach the adiabatic lapse-rate. Layers in which super-adiabatic lapse rates are present are inherently unstable, and convection and mixing will effect a redistribution heat energy such that the final lapse rates in these layers will not be super-adiabatic. Therefore, in those layers in which the computed lapse rates are super-adiabatic, the actual lapse-rates are probably close to adiabatic. In those layers in which the computed lapse rate is sub-adiabatic, and this includes most of the atmosphere between the surface and the cloud-base, the actual temperatures should be quite similar to the computed temperatures.

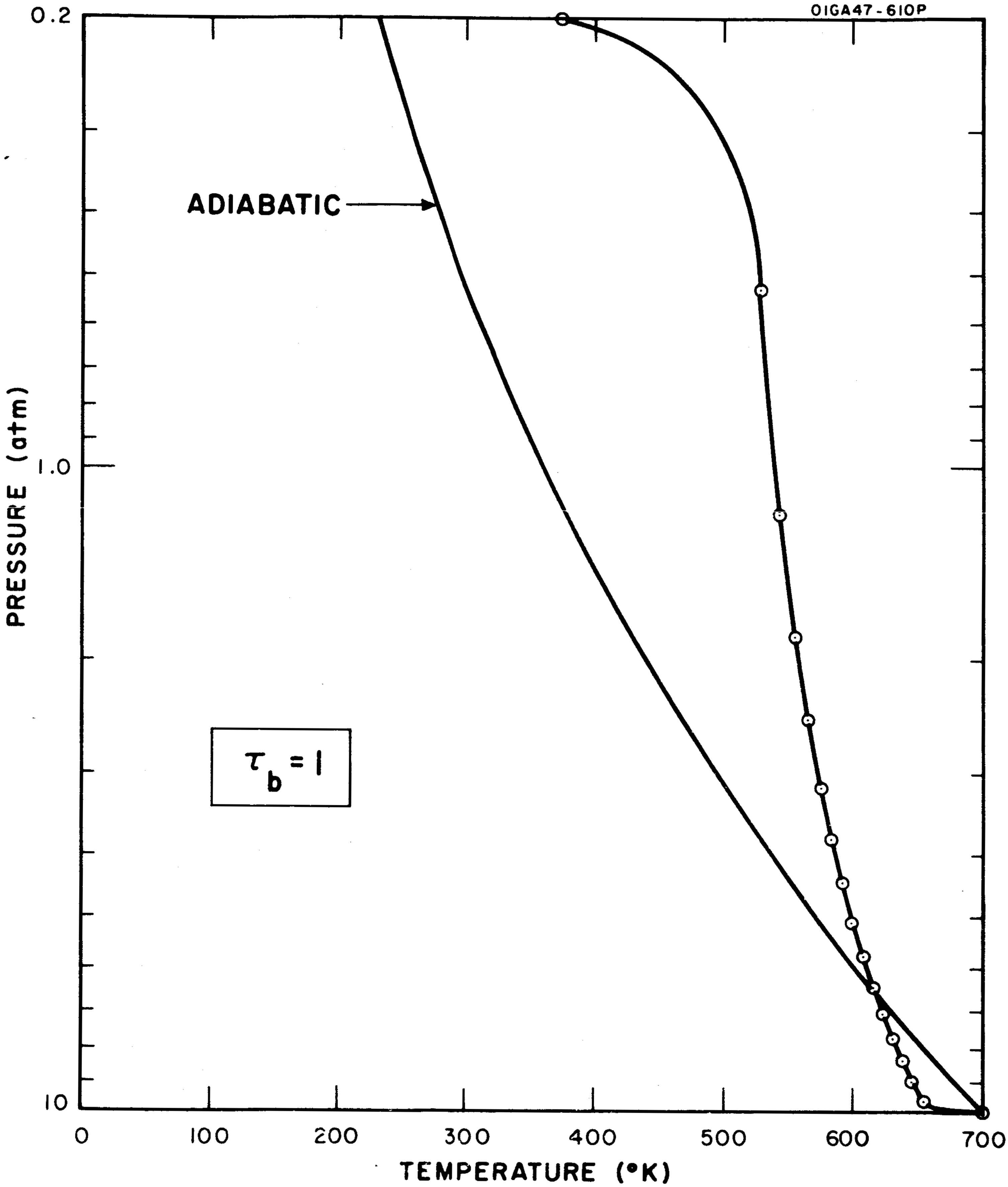

Figure 1. Computed radiative equilibrium temperature distribution between surface and cloud-base on Venus for atmospheric infrared opacity of 1. Adiabatic temperature distribution is also shown.

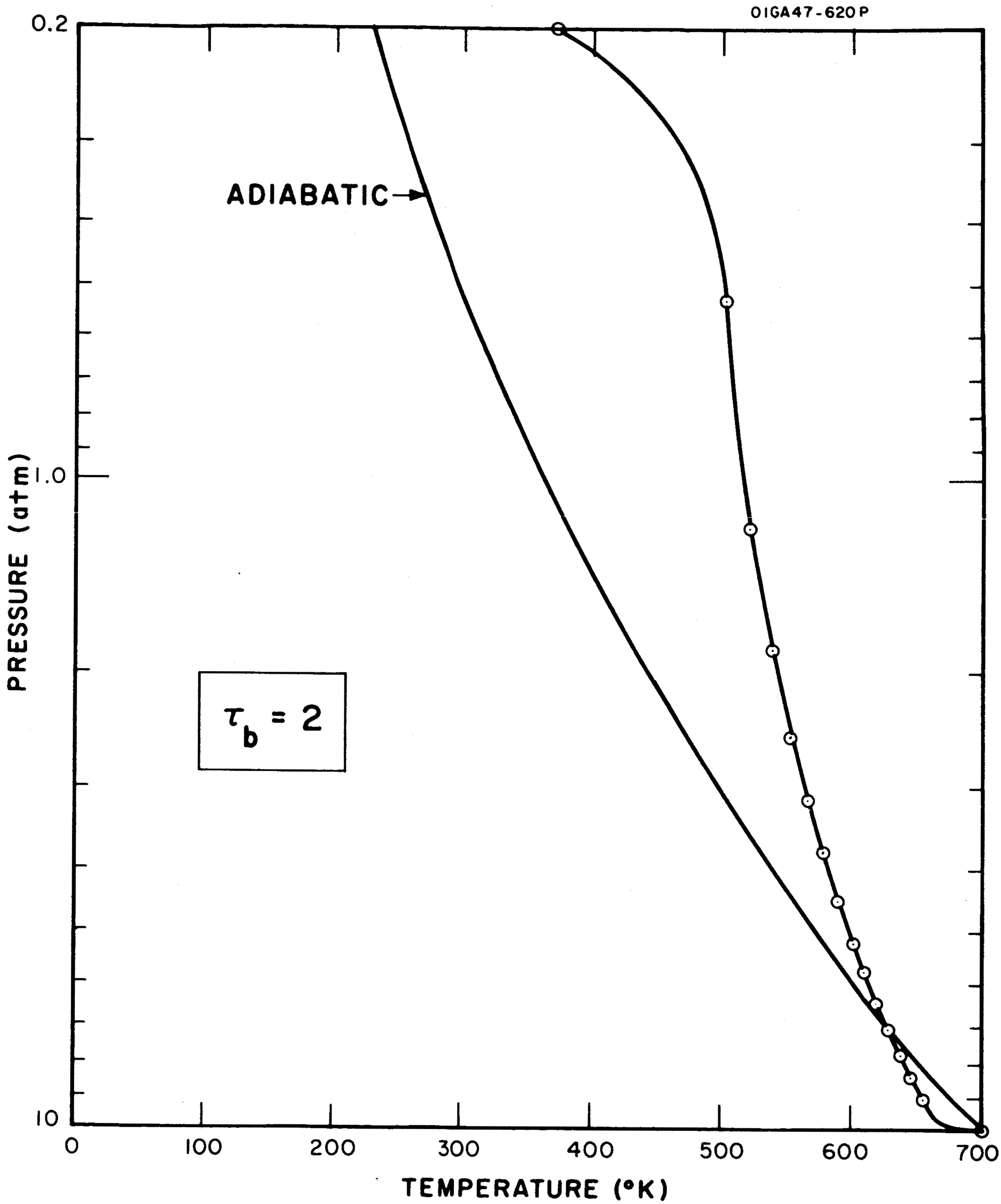

Figure 2. Computed radiative equilibrium temperature distribution between surface and cloud-base on Venus for atmospheric infrared opacity of 2. Adiabatic temperature distribution is also shown.

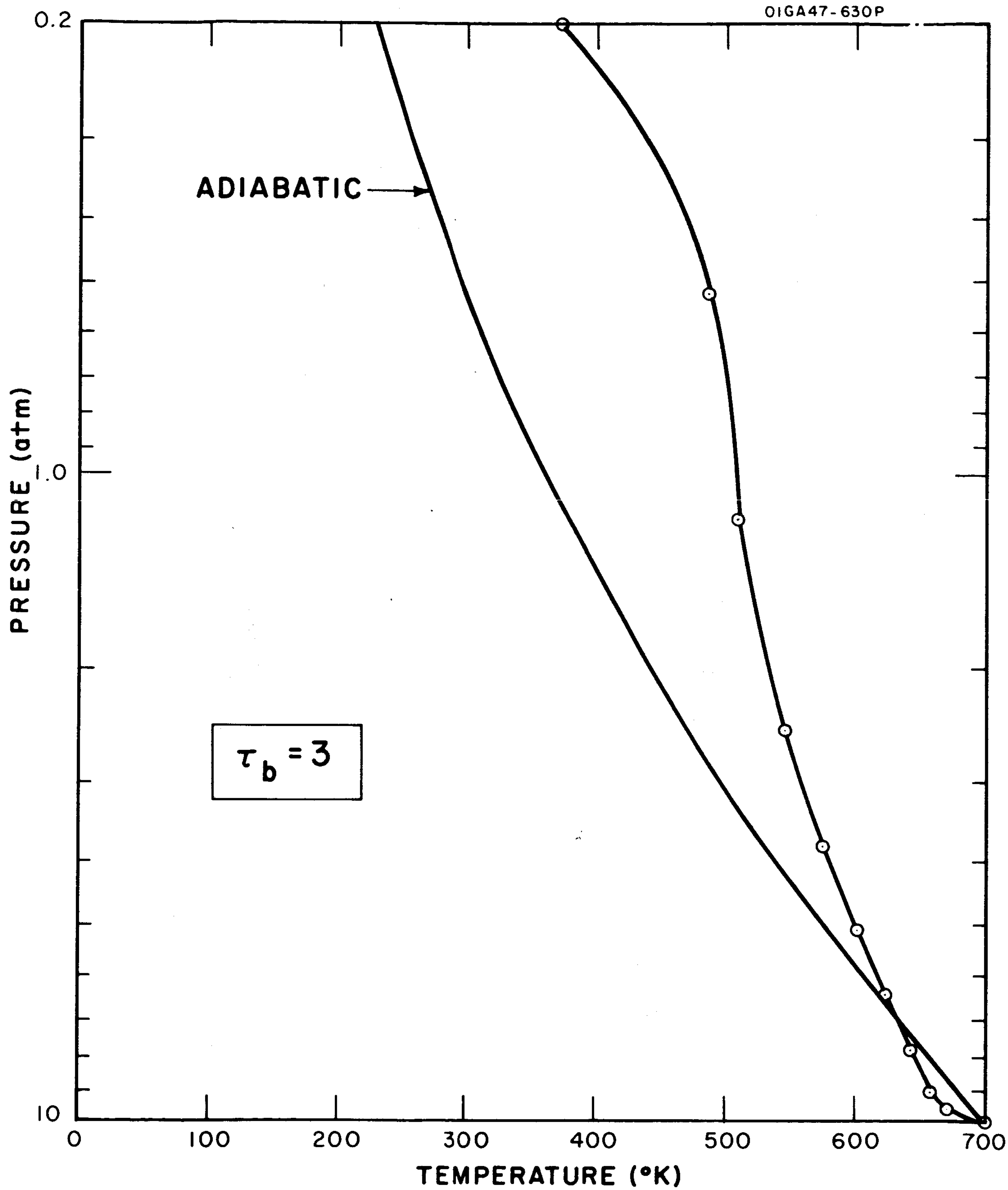

Figure 3. Computed radiative equilibrium temperature distribution between surface and cloud-base on Venus for atmospheric infrared opacity of 3. Adiabatic temperature distribution is also shown.

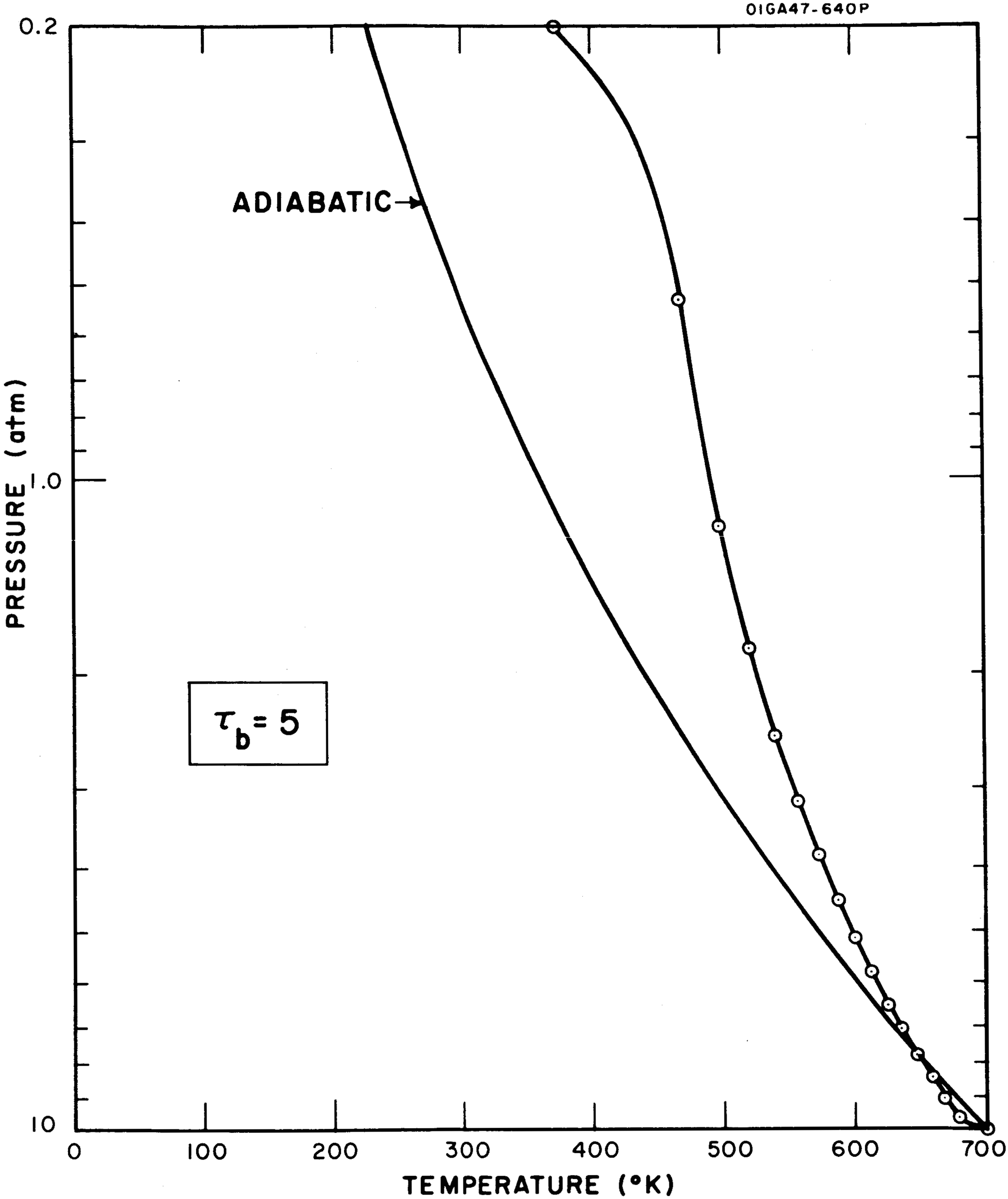

Figure 4. Computed radiative equilibrium temperature distribution between surface and cloud-base on Venus for atmospheric infrared opacity of 5. Adiabatic temperature distribution is also shown.

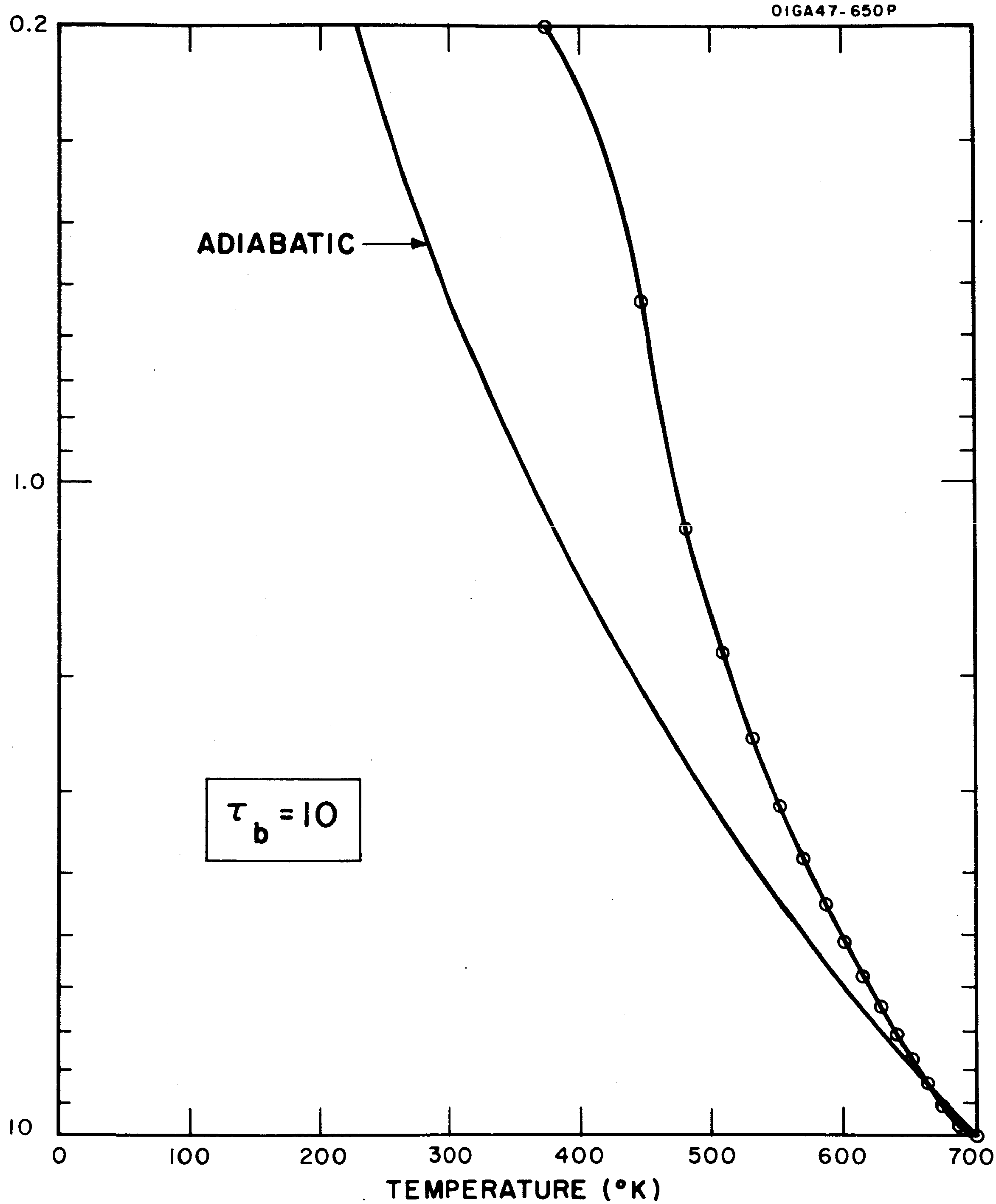

Figure 5. Computed radiative equilibrium temperature distribution between surface and cloud-base on Venus for atmospheric infrared opacity of 10. Adiabatic temperature distribution is also shown.

The actual value of the infrared opacity, τ_b, between the cloud-base and the surface of the planet is unknown. Computations of the greenhouse effect in the Venus atmosphere (Ohring and Mariano, 1964) indicate that, with 99% cloudiness, infrared opacities of 6 or 7 for the entire atmosphere are required to maintain a surface temperature of 700°K. Since the cloud base is located at 2×10^{-1} atm, which is 2×10^{-2} of the surface pressure, most of the atmosphere is below the cloud base. Consequently, the infrared opacity of the entire atmosphere is a fairly good value for the infrared opacity of the sub-cloud layer. Therefore, the actual temperature profile might best be represented by a profile intermediate between those computed for $\tau_b = 5$ and $\tau_b = 10$.

The radiative equilibrium temperature profiles computed in this study can be considered first approximations to the actual vertical temperature structure below the Venus cloud cover. The closeness of the approximation to the actual temperature structure depends upon a number of factors. These include the reality of the physical model used in the computations and the extent to which the atmosphere below the clouds is in infrared radiative equilibrium.

III. METEOROLOGY OF MARS

Over the course of a year, the amount of solar radiation received by a planet is balanced by the amount of infrared radiation emitted by the planet. However, at any one time, or at any one location, incoming solar radiation does not, in general, equal outgoing infrared radiation. The large scale variations of the difference between incoming and outgoing radiation represent heat energy sources and sinks; these sources and sinks are the driving forces of the general circulation of a planetary atmosphere, and knowledge of their magnitude and distribution is a basic requirement for an understanding of planetary circulations.

The net incoming solar radiation at the top of the Martian atmosphere is solely a function of the solar constant, solar-Martian geometry, and Martian planetary albedo. The solar constant and the required geometry are well known; the Martian planetary albedo, and, in particular, its variations with latitude and season are known to a lesser degree of accuracy. The outgoing radiation depends upon the surface temperatures, and the composition and temperatures of the atmosphere. Both composition and temperature, especially variations with latitude, altitude, and season, are not known very well; however, there is sufficient information available to allow construction of atmospheric models that can be used for preliminary estimates of the radiation budget. In the present examination of the problem, we shall attempt to obtain a preliminary estimate of the annual radiation budget of Mars - the average variation with latitude of incoming and outgoing radiation at the top of the Martian atmosphere.

The variation of incoming solar radiation with latitude at the top of the Martian atmosphere - before correction for albedo losses - was computed some time ago by Milankovitch (1920). These values are shown in column 2 of Table 2 and are based upon a solar constant of 2 cal cm^{-2} min^{-1} at the Earth's distance from the Sun. The variation of Martian planetary albedo with latitude has recently been estimated by Sagan (1964), and the average hemispheric variation is shown in Table 2. By multiplying the incoming solar radiation by (1 - A), where A is the albedo, one can obtain the latitudinal variation of net incoming radiation, which is the incoming part of the radiation budget.

To determine the outgoing radiation, knowledge is required of the mean latitudinal variation of surface temperature, of vertical distribution of temperature, and of atmospheric absorption characteristics. In the present model, the mean latitudinal variation of surface temperature is based upon Ohring et al's (1963) processing of Gifford's (1956) analysis of mid-day radiometric observations; this processing consisted of extrapolation of Gifford's temperature curves to the poles and correction for diurnal variations. These surface temperatures are shown in Table 2.

For the atmospheric temperatures and absorption characteristics, it is assumed that:

Table 2. Annual radiation budget of Mars.

Latitude deg	I_o cal $cm^{-2}min^{-1}$	A	$I_o(1 - A)$	T_g ^oK	F(o) cal $cm^{-2}min^{-1}$
0 - 10	0.262	0.200	0.209	238	0.194
10 - 20	.254	.185	.207	238	.194
20 - 30	.240	.178	.198	236	.187
30 - 40	.220	.185	.180	234	.181
40 - 50	.195	.180	.160	232	.175
50 - 60	.167	.160	.140	228	.163
60 - 70	.139	.150	.118	223	.149
70 - 80	.125	.165	.104	218	.136
80 - 90	.118	.185	.096	210	.118

1) The atmosphere consists of two layers: a troposphere, in which the temperature decreases with height at a rate of 0.9 times the adiabatic lapse rate, or about $3.3^{o}K/km$, and a stratosphere, in which the temperature remains constant with altitude.

2) The atmosphere acts as a grey absorber in the infrared region of the spectrum.

With these assumptions, the atmospheric temperature structure can be written as

$$T/T_g = (\tau/\tau_g)^{0.25} \tag{1}$$

where T is temperature, τ is infrared opacity increasing downward from the top of the atmosphere, and the subscript g refers to the planet's surface. The black-body flux B is equal to σT^4; therefore,

$$(B/B_g) = (\tau/\tau_g) \quad . \tag{2}$$

The outgoing flux of radiation at the top of the atmosphere can now be written as

$$F(o) = B_g \; 2E_3(\tau_g) + (2B_g/\tau_g)\int_{\tau_t}^{\tau_g} \tau \; E_2(\tau)\,d\tau + (2B_g\tau_t/\tau_g)\int_o^{\tau} E_2(\tau)\,d\tau. \tag{3}$$

Upon integration, this equation can be written as

$$F(o) = B_g \Big\{ 2E_3(\tau_g) + (2/3\tau_g)[\tau_t \; 2E_3(\tau_t) + e^{-\tau_t} - \tau_g \; 2E_3(\tau_g) - e^{-\tau_g}] + (\tau_t/\tau_g)[1 - 2E_3(\tau_t)] \Big\} \; . \tag{4}$$

If B_g, which depends solely on surface temperature, and τ_g and τ_t are known, the outgoing radiation can be computed from (4). It is assumed that τ_g and τ_t do not vary with latitude. The values of τ_g and τ_t to be used in (4) can be determined from the requirement that the average incoming solar radiation must equal the average outgoing radiation and the assumption that the stratosphere is, on the average, in gross infrared radiative equilibrium. The requirement can be written as

$$\bar{I}_o(1 - \bar{A}) = \bar{F}(o) = \sigma \bar{T}_g^4 \Big\{ 2E_3(\tau_g) + (2/3\tau_g)[\tau_t 2E_3(\tau_t) + e^{-\tau_t} - \tau_g 2E_3(\tau_g) - e^{-\tau_g}] + (\tau_t/\tau_g)[1 - 2E_3(\tau_t)] \Big\} \tag{5}$$

where $\bar{I}_o$ is the average incoming solar radiation at the top of the Martian atmosphere, $\bar{A}$ is the average Martian albedo, σ is the Stefan-Boltzmann constant, and $\bar{T}_g$ is the average surface temperature. The assumption can be written as

$$\bar{F}(o) = \bar{F}(\tau_t) \tag{6}$$

where $\bar{F}(o)$ is the average outgoing infrared radiation flux, and $\bar{F}(\tau_t)$ is the average infrared flux at the tropopause. For the present model, this requires that τ_g and τ_t be related through the following equation:

$$2E_3(\tau_g - \tau_t) + (2/\tau_g)\int_{\tau_t}^{\tau_g} \tau\, E_2(\tau - \tau_t)d\tau - 2(\tau_t/\tau_g)[1 - 2E_3(\tau_t)] = 2E_3(\tau_g) + (2/\tau_g)\int_{\tau_t}^{\tau_g} \tau\, E_2(\tau)d\tau \ . \tag{7}$$

The net incoming solar radiation can be written in terms of an effective black-body flux and temperature.

$$\overline{I}_o(1 - \overline{A}) = \overline{B}_e = \sigma \overline{T}_e^{\,4} . \qquad (8)$$

Given the ratio of $\overline{T}_g$, the observed average surface temperature, to $\overline{T}_e$, the effective temperature the surface would have in the absence of an atmosphere, it is possible to determine τ_g and τ_t from Equations (5) and (7). In Quarterly Progress Report No. 2, we performed the reverse computation; that is, given τ_g, we computed (T_g/T_e) and τ_t.

The average incoming solar radiation is 0.215 cal $cm^{-2}min^{-1}$; the average albedo, from the values in Table 2, is 0.18. Consequently, the value of $\overline{T}_e$ is 216^oK. The average surface temperature, from the values in Table 2, is 233^oK, so that the ratio (T_g/T_e) is equal to 1.08. With T_g/T_e = 1.08, we find, from Table II of Quarterly Progress Report No.2, that $\tau_g = 0.5$ and $\tau_t = 0.21$. Substituting these values into Equation (4), we find that

$$F(o) = .743\ B_g = .743\ \sigma\ T_g^{\,4} \qquad (9)$$

or, with $\sigma = 8.13 \times 10^{-11}$ cal cm^{-2} deg^{-4} min^{-1},

$$F(o) = 6.04 \times 10^{-11}\ T_g^{\,4} . \qquad (10)$$

Using Equation (10) and the surface temperatures listed in Table 2, we have computed the outgoing radiation flux as a function of latitude; these values are tabulated in the last column of Table 2. In Figure 6

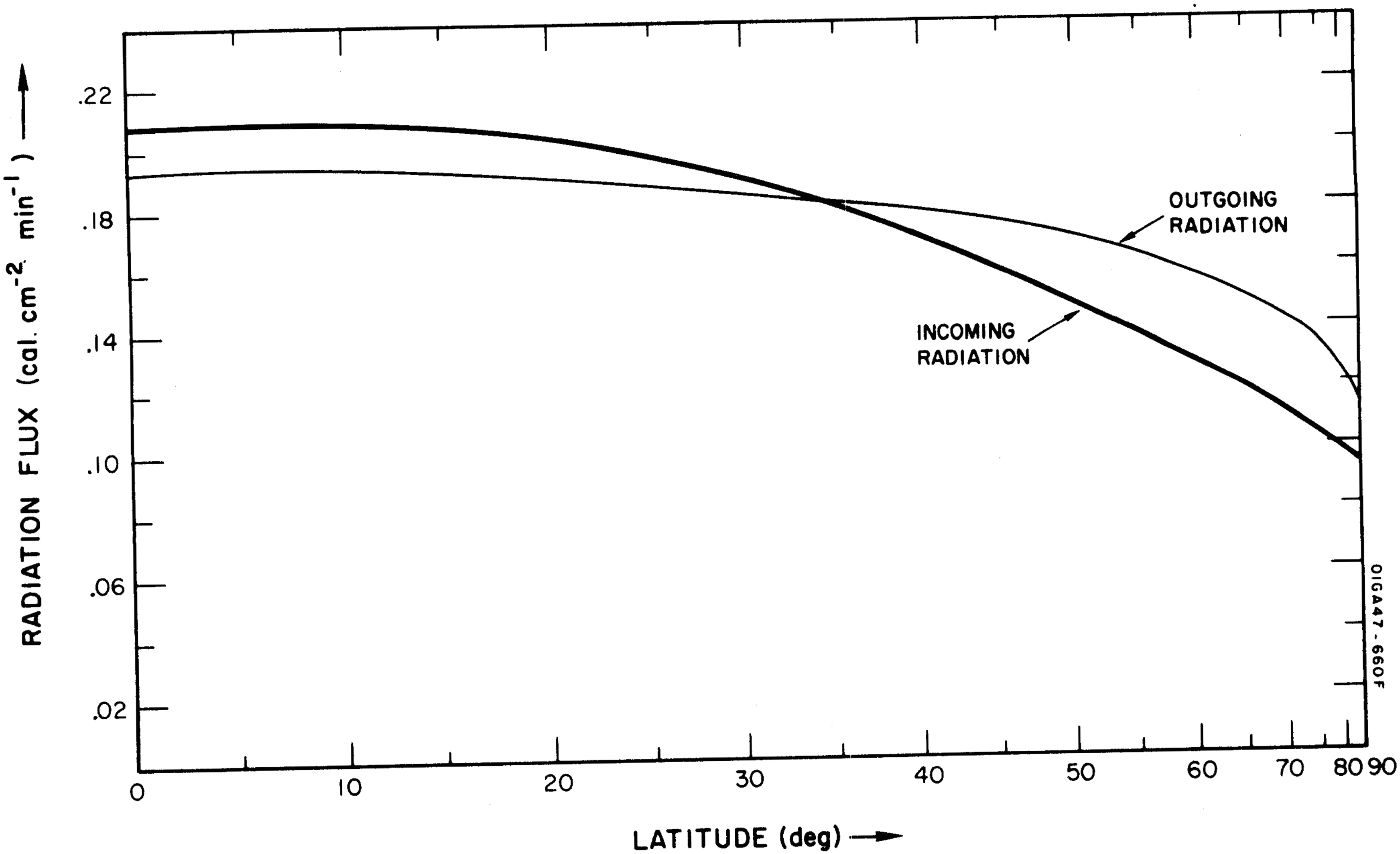

Figure 6. Annual radiation budget for planet Mars.

are plotted the variation with latitude of the net incoming solar radiation, $I_o(1 - A)$, and the outgoing radiation, $F(o)$; the latitude scale is a cosine scale, so that length is proportional to the area of a latitude belt. These curves represent the annual radiation budget of the planet Mars. It can be noted, from Figure 6, that below latitude 35^o there is a surplus of energy, and above 35^o there is a deficit of energy. These curves are quite similar to those computed for the Earth (See, for example, London, 1957); for the Earth, the cross-over point is at about 40^o latitude. Since the temperature does not rise from year to year at equatorial regions, nor fall at polar regions, the excess heat energy in equatorial regions must be transported to higher latitudes. The average rate of transport across a fictitious wall at the cross-over latitude can be determined from the area between the net incoming radiation curve and the outgoing radiation curve, below the cross-over latitude. This turns out to be about 0.5×10^{16} cal min^{-1}; for the Earth, the corresponding figure is about 5×10^{16} cal min^{-1} (after London, 1957). The Martian value is an order of magnitude less than the value for the Earth, and there are probably several reasons for this. The primary reason is probably the difference in surface area between the two planets; the amount of heat energy to be transported past the cross-over latitude depends upon the surface area of the planet below the cross-over latitude- and Mars' surface area is a factor of 4 less than the Earth's surface area. Another reason is probably the difference between Earth and Mars in the dependence of outgoing radiation with latitude. On Mars, the variation with latitude of the outgoing radiation depends essentially on the variation with latitude of the surface temperature; on Earth the latitudinal variation of outgoing radiation depends not only on the surface temperature variation but also on the water vapor variation with latitude. The latitudinal variation of water vapor in the Earth's atmosphere causes the latitudinal variation of outgoing radiation to be smaller than it would otherwise be. Hence, relatively larger differences between incoming and outgoing radiation occur, resulting in a larger value for the amount of excess heat energy.

With this information on the annual radiation budget, it is possible for dynamic meteorologists to infer the characteristics of the general circulation of a planet (See, for example, Mintz, 1961). We are currently studying these relationships. Besides the annual radiation budget, it is desirable to have seasonal radiation budgets; we have started work along these lines.

IV. ADMINISTRATIVE NOTES

The principal investigator presented a paper on Changes in the Amount of Cloudiness and the Average Surface Temperature of the Earth at the annual meeting of the American Meteorological Society, January 29-31, 1964, in Los Angeles. The greenhouse model upon which this paper was based was developed under the present contract.

The article The Effect of Cloudiness on a Greenhouse Model of the Venus Atmosphere, by George Ohring and Joseph Mariano, appeared in the January 1, 1964 issue of the Journal of Geophysical Research. The research reported on in this paper was performed under the present contract.

The American Meteorological Society has invited the principal investigator to write an Educational Monograph on the subject of The Atmospheres of the Planets. The Educational Monograph series of the American Meteorological Society, a series of short books on the atmospheric sciences, is designed for high school science students.

V. FUTURE PLANS

With the present contract scheduled to terminate on 4 March 1964, the preparation of the Final Report will occupy most of our time for the remainder of the contract.

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